Shaking intensity from injection-induced versus tectonic earthquakes in the central-eastern United States

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Abstract

Although instrumental recordings of earthquakes in the central and eastern United States (CEUS) remain sparse, the U. S. Geological Survey's "Did you feel it?" (DYFI) system now provides excellent characterization of shaking intensities caused by induced and tectonic earthquakes. Seventeen CEUS events are considered between 2013 and 2015. It is shown that for 15 events, observed intensities at epicentral distances greater than ≈ 10 km are lower than expected given a published intensity-prediction equation for the region. Using simple published relations among intensity, magnitude, and stress drop, the results suggest that 15 of the 17 events have low stress drop. For those 15 events, intensities within ≈ 10-km epicentral distance are closer to predicted values, which can be explained as a consequence of relatively shallow source depths. The results suggest that those 15 events, most of which occurred in areas where induced earthquakes have occurred previously, were likely induced. Although moderate injection-induced earthquakes in the central and eastern United States will be felt widely because of low regional attenuation, the damage from shallow earthquakes induced by injection will be more localized to event epicenters than shaking tectonic earthquakes, which tend to be somewhat deeper. Within approximately 10 km of the epicenter, intensities are generally commensurate with predicted levels expected for the event magnitude.

Introduction

Several studies have considered the hazard implications of induced earthquakes (see Ellsworth, 2013). One of the key issues for hazard assessment is the level of shaking generated by induced earthquakes. In standard probabilistic seismic hazard analysis (PSHA), ground motions for potential future sources are calculated using so-called ground-motion prediction equations (GMPE). GMPEs predict shaking intensity expected as a function of distance for a given magnitude. Some GMPEs consider faulting type and/or rupture directivity, but they generally rely on magnitude to characterize earthquake source size (see Cotton et al. [2013] for a recent review).

In recent decades, seismologists have adopted moment magnitude M_W as the best single measure of earthquake size. However, M_W is a static measure, depending only on the area of fault that moves, the average slip, and the rigidity of the surrounding crust. As discussed by Hough (2014), M_W does not reflect directly the radiated energy, which depends on the details of rupture. A second parameter, stress drop, describes to first order the spectrum of radiated energy for a given M_W. Although stress drop is known to vary significantly among earthquakes, the parameter is difficult to estimate and is not included as a variable in modern GMPEs.

In a recent study, Hough (2014) considers the shaking generated by 11 M_W 3.9–5.7 earthquakes that previously published studies identified as likely to have been induced by fluid injection (Figure 1). As discussed by Hough (2014), although they are not determined uniformly, moment-magnitude estimates are available for all these inferred induced events, and inconsistency in magnitude determination cannot account plausibly for observed systematics. Spatially rich, intensity data sets also are available for all these events from the USGS "Did you feel it?" system (Wald et al., 1999).

There is a growing appreciation for the utility of spatially rich, systematically determined DYFI data to address key questions in earthquake ground-motion science. Intensity data long have been used to investigate historical earthquakes for which instrumental data are not available. Even in modern times, compared with the relatively limited number of instrumental ground-motion recordings of a given earthquake, community decimal intensity (CDI) values calculated systematically from responses to the DYFI system provide more spatially rich sampling.

Although intensities historically have been reported with Roman numerals, CDI values are reported as decimal values to one-tenth unit. Although not an instrumental measure of ground motions, DYFI intensities serve as reliable proxies for ground-motion parameters such as peak ground acceleration (PGA) (Atkinson and Wald, 2007). The DYFI system calculates an intensity from every submitted response within a given ZIP code. When multiple responses are submitted within a

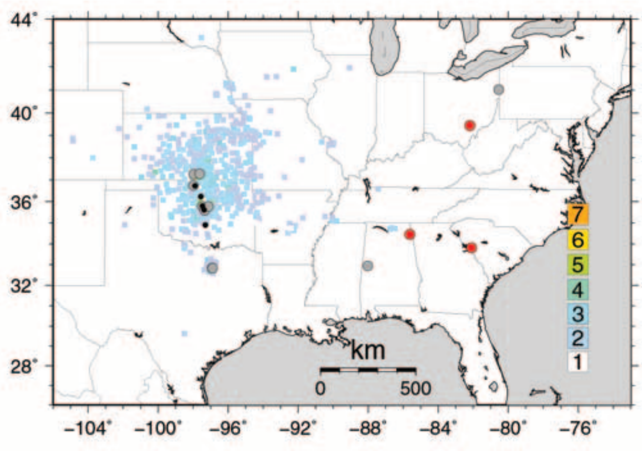

Figure 1. Locations of the 17 earthquakes analyzed in this study, including 10 events listed in Table 1 that I suggest were induced (larger gray dots) and five events in Table 2, all of which are suggested to be induced (small black dots). The three inferred tectonic events are indicated by red circles. The DYFI reported intensity values (in numbered legend to the right) are shown for the inferred induced M 4.9 earthquake in Kansas on 12 November 2014.

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Table 1. Earthquakes analyzed in this study: year, month, day, magnitude (M), magnitude type, depth in km (z), estimated M_{IE}, number of DYFI responses (N_R), latitude, longitude, and state. Magnitudes and locations are reported from the National Earthquake Information Center, with magnitude type indicated: moment magnitude (M_W, generally so-called regional M_W estimates), body-wave magnitude (M_b), duration magnitude (M_d). Source depths of 5 km are assigned, indicating a poorly constrained depth estimate.

Year	Month	Day	M	Type	z	M_{IE}	N_R	Lat.	Long.	State
2003	4	29	4.6	M_L	9.1	4.8	17600	34.445	−85.620	AL
2013	11	20	3.5	M_b	7.9	3.3	465	39.439	−82.190	OH
2014	2	15	4.1	M_d	5.1	4.3	15676	33.817	−82.092	SC
2014	2	17	3.8	M_W	7.4	3.0	563	35.776	−97.469	OK
2014	3	10	3.0	M_b	2.5	2.4	43	41.010	−80.543	OH
2014	6	16	4.3	M_W	5.0	3.5	1041	35.577	−97.340	OK
2014	7	29	4.3	M_W	7.6	3.9	1367	36.732	−97.987	OK
2014	10	2	4.3	M_W	5.0	3.7	1267	37.245	−97.955	KS
2014	10	10	4.2	M_W	5.0	3.6	901	35.748	−97.123	OK
2014	11	12	4.9	M_W	4.0	4.0	5343	37.271	−97.621	KS
2014	11	20	3.8	M_W	5.0	2.5	55	32.950	−88.017	AL
2015	1	6	3.5	M_b	4.2	2.8	2772	32.836	−96.901	TX
2015	1	7	3.6	M_b	5.0	2.7	1860	32.837	−96.890	TX

Table 2. M ≈ 4 earthquakes in Oklahoma analyzed in this study: year, month, day, magnitude, magnitude type, depth in kilometers (z), number of DYFI responses (N_R), estimated M_{IE}, latitude, and longitude. Magnitudes and locations are reported from the National Earthquake Information Center, with magnitude type indicated: moment magnitude (M_W) and body-wave magnitude (M_b). Source depths of 5 km are assigned, and they generally indicate a poorly constrained depth estimate.

Year	Month	Day	M	Type	z	N_R	M_{IE}	Lat.	Long.
2014	2	9	4.1	M_W	5.0	414	3.3	34.893	−97.292
2014	4	10	4.0	M_b	3.4	489	3.4	35.791	−97.471
2014	6	18	4.1	M_W	5.0	309	3.0	35.610	−97.371
2014	7	15	3.9	M_W	5.0	279	3.2	36.713	−97.888
2014	9	30	4.0	M_W	2.2	201	3.1	36.224	−97.554

ZIP code, the individual intensity assignments are averaged. By definition, reported DYFI intensities reflect representative rather than extreme intensities within the spatial footprint of a ZIP code.

Hough (2014) shows that within ≈ 10-km epicentral distance, intensities for the 11 inferred induced events are generally commensurate with predictions from published intensity-prediction equations, but at greater distances, intensities are systematically lower. Hough (2014) introduces an effective intensity magnitude M_{IE}, defined to be the magnitude that best fits an observed intensity distribution, given an existing intensity-prediction equation for the central and eastern United States (CEUS). On average, for the events analyzed, M_{IE} was lower than M_W by an average of 0.8 units for the 11 earthquakes believed to be induced. In contrast, for a set of 10 inferred tectonic earthquakes, Hough (2014) finds M_{IE} to be comparable to M_W. It follows, then, that DYFI data are potentially useful to identify earthquakes that might have been induced, meriting further study.

In this study, I analyze an additional 17 events, including (1) all the CEUS earthquakes since 2013 for which at least 500 DYFI responses were received, as well as three recent smaller earthquakes in Alabama and Ohio, and (2) DYFI data for an additional five M 3.9–4.1 events in Oklahoma for which fewer than 500 but more than 200 DYFI responses were received (Figure 1). Although fewer DYFI data are available for the second set of events, trends in CDI versus distance data can be resolved using stacked observations from multiple events. In contrast to Hough (2014), for this study, I analyzed all events that fit the above criteria rather than selecting a data set of events identified previously as having been either induced or tectonic. Definitive analyses have not been published yet for many of the events analyzed in this study; it is thus not known in advance which events are likely to be induced. Magnitude estimates have not been determined consistently for all events, but M_W values are available for 11 of the 17 events.

Analysis of recent events

I first compare DYFI data for the 12 events listed in Table 1 with predicted intensities from relationships developed to fit DYFI data from the CEUS (Figure 2). Hough (2014) shows that the curves developed by Atkinson and Wald (2007) provide a generally good fit to data from moderate tectonic events. I use the Atkinson-Wald relations here because they were developed using data from events prior to 2007, that is, before the increase in CEUS seismicity associated with the increase in injection-induced events (Ellsworth, 2013). Community-decimal-intensity values from the DYFI system are fit by intensity-prediction relationships that include a nonlinear magnitude term as well as a piecewise-continuous distance decay:

$$\mathrm{CDI}\,(M_W, R) = d_1 + d_2(M_W-6) + d_3(M_W-6)^2 + d_4\log(R) + d_5 R + d_6 B + d_7 M_W \log(R), \tag{1}$$

in which

$$R = \sqrt{(D^2 + h^2)},$$

and

$$B = 0 \text{ for } D \leq D_t;$$
$$B = \log (D/D_t),\ D > D_t.$$

Here, d_1 through d_7 are constants, and D_t is a transition distance that the Atkinson-Wald estimate to be 80 km for CEUS earthquakes. The parameter D is defined to be the nearest distance to the fault, which in theory is equivalent to hypocentral distance for small to moderate events. The parameter h is introduced to stabilize the inversion and can be regarded as an effective depth. Although R is thus a nonphysical parameter, because the bulk of the DYFI data for all events in this study is from distances greater than 20 km, R is effectively comparable to hypocentral distance. I will return to the question of near-field intensities in a later section.

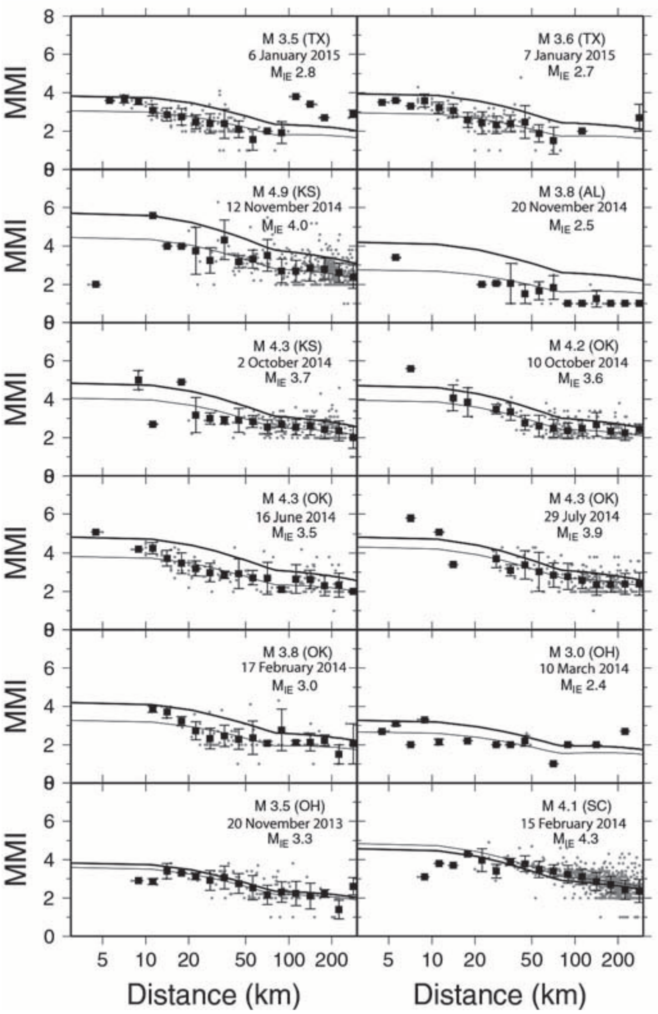

Figure 2. The DYFI intensities for 12 of the events listed in Table 1, with average values in logarithmic hypocentral distance bins, ± 1 standard deviation (black squares). Dark line indicates predicted intensity curve using Atkinson and Wald (2007) CEUS relations given estimated moment magnitudes.

Figure 2 shows intensity prediction curves for each instrumentally determined magnitude; for all analysis, I use the d_1 through d_7 and h and D_t values determined for the CEUS by Atkinson and Wald (2007). As observed by Hough (2014), all the data sets reveal the bias associated with underreporting ZIP codes; i.e., there is a tendency of CDI(R) to flatten at the largest distances and an absence of CDI values between 1.0 (which corresponds to no felt reports received) and 2.0 (which is the minimum CDI assigned if even a single report of felt shaking is received within a ZIP code). Because the Atkinson-Wald relationships were developed using DYFI data that share this same bias, it is arguably appropriate to compare the data from all distances with predicted intensities. For this study, however, I focus on data within a distance of 100 km because these are considered more reliable.

Following Hough (2014), I calculate the magnitude value M_{IE}, which optimizes the least-squares fit between the intensities and equation 1, considering data from distances within 100 km. Given the optimal M_{IE} values, predicted intensities from equation 1 are generally consistent with bin-averaged intensity values ± 1 standard deviation. Intensities generally are closer to predictions for the event magnitude for distances within ≈ 10 km.

The DYFI data generally are more sparse and less reliable for smaller events. I also note that if intensities for induced earthquakes are generally low, they will be felt less strongly than tectonic earthquakes of comparable magnitude, and therefore they will generate smaller DYFI data sets. As an illustrative example, one can consider the DYFI response to two events with essentially the same magnitude, both of which occurred not far from urban centers. Only 55 DYFI responses were received for the M 3.8 Alabama earthquake of 20 November 2014 (Table 1), which I suggest was likely an induced event. However, 646 responses were received for the M 3.9 Alabama earthquake of 12 January 2001 (Table 1), which I suggest was likely tectonic. If anything, responses are expected to increase over time from 1999 (when the DYFI system first was developed) to recent years as awareness of the Web site has grown. The 20 November 2014 event is included here because it occurred in a region with low background seismicity, raising the question of whether or not it was induced. The earthquake occurred in proximity to the Black Warrior Basin, where hydraulic-fracturing methods have been used increasingly in recent years (e.g., EPA, 2004). Data for the five M 3.9–4.1 events listed in Table 2 are shown in Figure 3, along with one panel showing combined data from all 17 events.

Interpretation

Of the 17 events analyzed, two have M_{IE} values that are within 0.2 magnitude units of the event magnitude: the 20 November 2013 M 3.5 event in Ohio and the 15 February 2014 M 4.1 event in South Carolina. For the other 15 events, the estimated M_{IE} value for all events is lower than the event magnitude by 0.4 to 1.3 magnitude units (with an average difference of 0.78 units), corresponding to a factor of 4–90 in seismic moment. These results are indistinguishable from the results of Hough (2014), who finds an average difference of 0.82 magnitude units, with the same range of (M_W–M_{IE}) values. In other words, the shaking levels generated by 15 of the 17 events analyzed in this study are commensurate with expected shaking for earthquakes 0.4 to 1.3 units lower than the event magnitude. As discussed by Hough (2014), it is implausible that the results are from inconsistency in magnitude estimation. The 15 events have not been analyzed as thoroughly by other authors as the events considered by Hough (2014), so it is not established which events are likely to be induced. If one considers the intensity data, in light of the results of Hough (2014) and event locations, it seems clear that 15 of the 17 events analyzed in this study likely were induced, and the two aforementioned earthquakes were tectonic. Almost all these 15 events occurred in regions where induced earthquakes have been documented previously. Notably, a recent study concluded that the M 3.0 Ohio earthquake on 10 March 2014 was induced by hydraulic fracturing (Skoumal et al., 2015). Analysis of additional events believed to have been induced by hydraulic fracturing will be needed to determine whether these events, like injection-induced earthquakes, also have generally low M_{IE} values. I note that the hydraulic-fracturing process per se is believed to induce relatively minor seismicity (e.g., Ellsworth, 2013).

In general, intensity distributions are assumed to be controlled primarily by three factors: magnitude, geometric spreading, and regional attenuation (e.g., Atkinson and Wald, 2007). Hough (2014) argues that one cannot appeal to attenuation differences to explain the systematic differences between the events analyzed here and the Atkinson-Wald intensity-prediction equations for the CEUS because for all events, the distance decay of DYFI intensity values was found to be consistent with equation 1. This result holds for the events analyzed in this study as well. Effectively, this indicates that regional attenuation of perceptible ground motions is comparable for injection-induced and tectonic earthquakes. Although one might conjecture that attenuation is locally higher in the vicinity of the induced events, that would lead to lower intensity values at close distances, which is contrary to the observations. Locally higher attenuation also could not explain why the distance decay is consistent with established regional intensity-prediction equations.

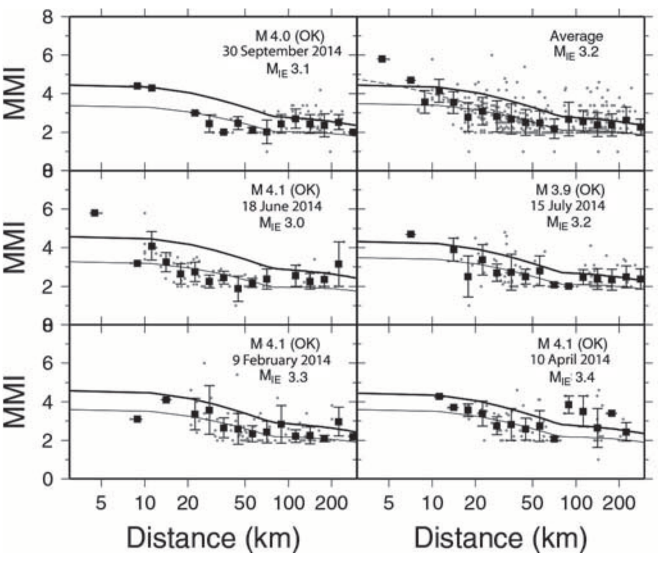

Figure 3. The DYFI Intensities for the events listed in Table 2 (gray dots), with average values in logarithmic hypocentral distance bins, ± 1 standard deviation (black squares). The top right panel shows combined data for all events. In each panel, dark and light lines, respectively, indicate predicted intensity curve using Atkinson and Wald (2007) CEUS relations for the event magnitude and the magnitude that provides the optimal fit to the Atkinson-Wald relation. In the top right panel, the dashed line indicates the predicted curve for the average M_{IE} value, assuming an effective depth of 3 km.

The new data set analyzed here provides the opportunity to compare directly intensities for two inferred induced earthquakes with DYFI data for nearby events that are known or believed to be tectonic (M. Brudzinski, personal communication, 2015; S. Jaume, personal communication, 2015). The comparison for one pair of events is shown in Figure 2: the 10 March 2014 inferred induced earthquake in Ohio (M 3.0; z = 2.5 km) versus the 20 November 2013 inferred tectonic event (M 3.5; z = 7.9 km). The former event occurred in eastern Ohio and the latter in southeastern Ohio. Figure 4 shows the comparison for a second pair of events: the 20 November 2014 inferred induced event in Alabama (M 3.8; z = 5 km) versus the 29 April 2003 tectonic earthquake in eastern Alabama (M 4.6; z = 9.1 km). Note that a reported depth of 5 km indicates a poorly resolved hypocentral depth. For the 2003 tectonic earthquake in Alabama, M_{IE} is higher than the instrumentally determined magnitude (4.8). For both pairs of events, the intensity data for the earthquakes inferred to be induced reveal a different signature even though the locations are not far apart and, to the extent that they are constrained, the depths of all events are relatively shallow.

As argued by Hough (2014), a systematic amplitude bias for induced earthquakes points to a significant difference in source properties. Considering basic scaling relationships, Hanks and Johnston (1992) show that high-frequency ground motions depend only weakly on M_W but strongly on stress drop. Using results from random vibration theory, peak acceleration and velocity, respectively, a_{max} and V_{max}, can be related to magnitude and stress drop σ:

$$\log a_{max} \approx 0.31\mathrm{M} + 0.80 \log(\sigma), \quad (2)$$

$$\log V_{max} \approx 0.55\mathrm{M} + 0.64 \log(\sigma). \quad (3)$$

These relations are derived using numerical approximations that are strictly valid for magnitudes greater than 5. This result is illustrated in Figure 5, which shows both theoretical velocity spectra for a range of magnitudes and a given stress drop and spectra for a range of stress-drop values for a given magnitude, assuming $\sigma = Mo(f_c/0.42\beta)^3$, where β is the shear-wave velocity near the source and f_c is the corner frequency. Figure 5 illustrates how strongly radiated energy for a given M_W, which is expected to depend on velocity squared, depends on stress drop.

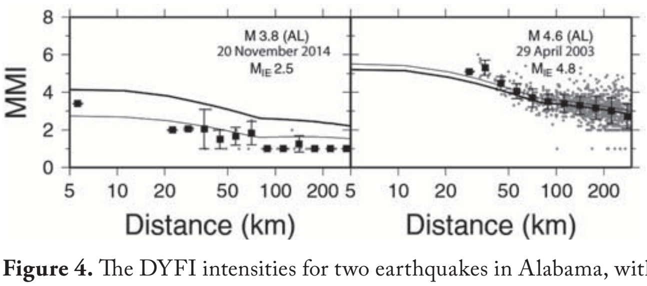

Figure 4. The DYFI intensities for two earthquakes in Alabama, with average values in logarithmic hypocentral distance bins and ± 1 standard deviation (black squares). In each panel, dark and light lines, respectively, indicate predicted intensity curve using Atkinson and Wald (2007) CEUS relations for the event magnitude and the magnitude that provides the optimal fit to the Atkinson-Wald relation.

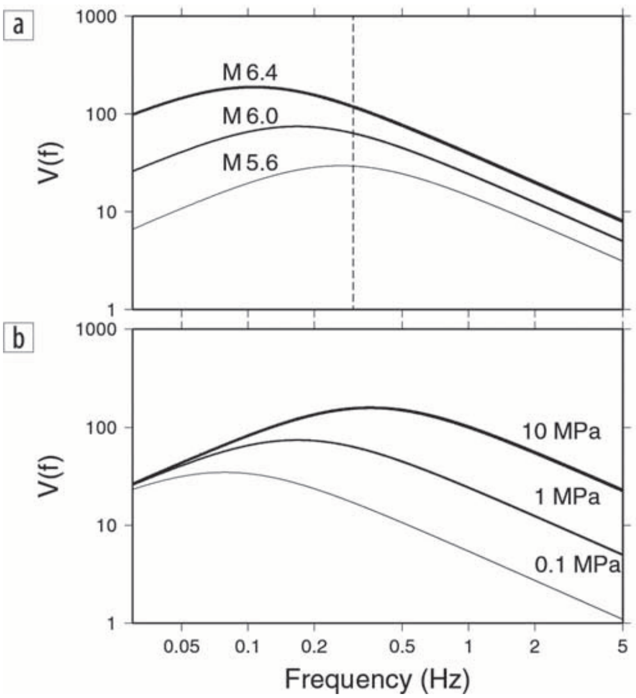

Figure 5. (a) A given stress-drop value and a range of magnitudes versus (b) theoretical (omega-squared) normalized source spectra for a given magnitude and a range of stress-drop values.

Assuming that the low to moderate intensities analyzed in this study are controlled by peak acceleration and that (M_W–M_{IE}) is controlled by source rather than path effects, one can use equation 2 to estimate the reduction in stress drop associated with a given value of M_{IE}: $10^{(-.39(M-M_{IE}))}$. Using equation 2, the inferred M_{IE} values correspond to a factor of 1.5–3.2 reduction in stress drop, for a given M_W. Alternatively, if intensities are controlled by peak velocity, a given value of (M_W–M_{IE}) corresponds to a stronger reduction in stress drop: 2.2 to 13. It remains unclear whether intensities are controlled more by peak acceleration or peak velocity; moreover, equations 2 and 3 are only approximations based on random vibration theory. The range and average (M_W–M_{IE}) values estimated in this study are identical to those inferred by Hough (2014), suggesting that stress drops for the 15 inferred induced events are lower by factors of ≈ 2–10 than average stress-drop values in the region. As discussed by Hough (2014), one interpretation is that induced earthquakes have lower stress-drop values because they are shallow. It is thus possible that DYFI data fundamentally provide an indication of source depth, although I note that all the events analyzed in this study were relatively shallow.

The inferred characteristic signature of intensities of induced earthquakes suggests that intensity data potentially can provide a first-order discriminant between induced and tectonic earthquakes. Hough (2014) discusses how the results of that study bear on the question of whether the second and third principal events in the Prague, Oklahoma, sequence were injection induced or tectonic.

In this study, analysis of DYFI data suggests that 15 of the 17 events analyzed were likely induced, and the 20 November 2013 M3.5 and 15 February 2014 M4.1 in Ohio and South Carolina, respectively, were tectonic.

As noted, although the bulk of DYFI data for all inferred events analyzed by this and the earlier study are more consistent with the inferred M_{IE} values than the event magnitudes, available intensity data for near-field distances, within 10 km, are more consistent with predictions for each event magnitude. Because, as noted, equation 1 includes nonphysical depth terms, there is a potential disconnect between these distances provided by the DYFI system and the distance term R in equation 1. The difference between distance measures will be consequential only for distances less than ≈ 20 km, distance ranges for which there are relatively little data in the CEUS, for either the Atkinson-Wald calibration events or the events analyzed in this study. However, the intensity-prediction equations determined using equation 1 provide a good fit to near- and far-field DYFI intensities for tectonic events. The question of interest is thus whether near-field DYFI intensities for injection-induced earthquakes differ from near-field intensities for tectonic events.

Relatively high near-field intensities for the induced earthquakes analyzed in this study are consistent with expectations for shallow events. Although depth cannot be estimated reliably from intensity data, basic wave-propagation considerations predict that shallow earthquakes will generate higher intensities in the epicentral region than deeper events. Considering the combined DYFI data for the M ≈ 4 events in Oklahoma (Table 2), except for one near-in data point controlled by limited observations, intensities at all distances are well fit by equation 1 given M_{IE} 3.2 and a hypocentral depth of 3 km.

I note another potentially important point regarding near-field intensities. As summarized by Hough (2014), intensity values tend to be distributed normally within the footprint of a large city, with outlier values commonly exceeding the average by 1 intensity unit and not uncommonly by 1.5 to 2 units. It is therefore reasonable to expect that intensities also will be distributed normally within the footprint of an individual ZIP code. Thus, for example, for M_{IE} 4.0 and a hypocentral depth of 5 km, equation 1 predicts intensities of 5.8 and 5.0 for epicentral distances of 1 and 10 km, respectively, but shaking effects at individual locations/structures are expected to be as much as 1.5 to 2 units higher. Thus it would be in keeping with expectations, for example, that an induced earthquake with M_W 4.8 (and M_{IE} 4.0) will generate isolated instances of damage commensurate with intensity of 6 or 7. Indeed, CDI values as high as 6 have been determined in the near field of a number of relatively modest ($4.0 \le M_W \le 4.5$) presumed injection-induced earthquakes in Oklahoma (USGS, 2015).

Conclusions

Although instrumental recordings of injection-induced earthquakes remain sparse in the central and eastern United States, the DYFI system now provides excellent characterization of shaking intensities caused by induced earthquakes. In this study, I consider 17 CEUS events between 2013 and 2015 and show that, for 15 events, DYFI intensities are consistent with effective intensity magnitudes that are lower by 0.4 to 1.3 units than instrumentally determined magnitudes, with an average difference of 0.8 units. Using simple relations among peak acceleration, magnitude, and stress drop inferred from standard scaling relations and random vibration theory, these factors suggest that stress-drop values for 15 of the 17 events are lower by factors of approximately 2 to 10 than average stress drop in the region. The 15 events also are shallow, with estimated depths of 2.5 to 7.6 km. The interpretation that low M_{IE} values indicate low stress drop can be tested further by comparing inferences based on DYFI data with stress-drop values estimated using conventional methods. At close distances, the reduction in shaking intensity from inferred lower stress drop appears to be offset by increases in intensities because of shallow depth. Although these are separate effects, they appear to approximately compensate each other in the near field. At distances beyond 10 km, I conclude that low stress drop is the primary control on shaking intensities. One cannot rule out the possibility that stress drops are lower because of the relatively shallow hypocentral depths, although this is not the preferred interpretation. The results of this study suggest that 15 of the 17 events, most of which occurred in areas where induced earthquakes have been documented in recent years, are induced.

Although moderate injection-induced earthquakes in the central and eastern United States will be felt widely because of low regional attenuation, the damage from earthquakes induced by injection will be more localized to the event epicenters than damage from tectonic earthquake because of lower stress drops of induced events. Regardless of the interpretation, a growing body of well-constrained DYFI data provides prima facie evidence that shaking from injection-induced earthquakes is significantly lower at regional distances than shaking from tectonic earthquakes in the same region. Within approximately 10 km of the epicenter, intensities generally are commensurate with levels expected for the event magnitude.

Data and resources

All intensity data are downloaded from the USGS "Did You Feel It?" Web site, http://earthquake.usgs.gov/earthquakes/dyfi, accessed 1/17/2015. Magnitudes and locations are from the ANSS Comprehensive Catalog (http://earthquake.usgs.gov/earthquakes/map/doc_aboutdata.php, accessed 1/17/2015.) Near field intensities for the 6 November 2011 Prague, Oklahoma, earthquake were provided by C. Frohlich (personal communication, 2015). TLE

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